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Quaternary glaciation and hydrologic variation in the South American tropics as reconstructed from the Lake Titicaca drilling project

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Abstract

A 136-m-long drill core of sediments was recovered from tropical high-altitude Lake Titicaca, Bolivia-Peru, enabling a reconstruction of past climate that spans four cycles of regional glacial advance and retreat and that is estimated to extend continuously over the last 370,000 yr. Within the errors of the age model, the periods of regional glacial advance and retreat are concordant respectively with global glacial and interglacial stages. Periods of ice advance in the southern tropical Andes generally were periods of positive water balance, as evidenced by deeper and fresher conditions in Lake Titicaca. Conversely, reduced glaciation occurred during periods of negative water balance and shallow closed-basin conditions in the lake. The apparent coincidence of positive water balance of Lake Titicaca and glacial growth in the adjacent Andes with Northern Hemisphere ice sheet expansion implies that regional water balance and glacial mass balance are strongly influenced by global-scale temperature changes, as well as by precessional forcing of the South American summer monsoon.

Keywords: paleoclimate, Lake Titicaca, Andes, Altiplano, Bolivia, Peru, South America, quaternary, diatoms, isotopes, glaciation

Introduction

The advance and retreat of continental ice sheets in the Northern Hemisphere high latitudes is a fundamental feature of the Quaternary period. Northern Hemisphere ice sheet expansion coincided with decreased concentrations of atmospheric CO₂ (Petit *et al.*, 1999) and decreased surface temperatures of global extent. In the southern tropics of South America, snow lines were lower during the last glacial maximum (LGM) (Broecker and Denton, 1989 and Klein *et al.*, 1999), about 20,000 to 25,000 yr BP, although the magnitude of temperature depression and precipitation change and the exact timing of regional glacial expansion and contraction relative to the northern continental ice sheet are still debated (Hostetler and Clark, 2000, Kull and Grosjean, 2000, Pierrehumbert, 1999, Porter, 2001, Seltzer *et al.*, 2002, and Smith *et al.*, 2005b). High lake levels document high precipi-

tation at the LGM in the southern tropical Andes, but precipitation patterns in the adjoining Amazon are less clear (Anhuf *et al.*, 2006, Mayle *et al.*, 2004, and Vizy and Cook, 2005). Prior to the LGM, the history of tropical glacial activity and hydrologic variation is very poorly constrained, because there are few long paleoclimatic records from the Southern Hemisphere tropics. As a result, the relative importance of global-scale glacial boundary conditions (Garreaud *et al.*, 2003) versus seasonal insolation (Clement *et al.*, 2004 and Martin *et al.*, 1997) in forcing glacial and hydrologic mass balances is unclear. Here we report the results of a major drilling project designed to reconstruct the history of glaciation and hydrologic variability in the tropical Andes of South America from the sedimentary record of Lake Titicaca, Bolivia/Peru. Our results complement the only other long drill core records of terrestrial climate in tropical South America, namely those of Van der Hammen and Hooghiemstra from the northern tropical An-

des of Columbia (Van der Hammen and Hooghiemstra, 2000 and Van't Veer and Hooghiemstra, 2000) and our own previous studies from the Salar de Uyuni, farther south on the Bolivian Altiplano (Baker *et al.*, 2001a and Fritz *et al.*, 2004), as well as the long speleothem records from Brazil (Cruz *et al.*, 2005 and Wang *et al.*, 2004).

Environmental setting

Lake Titicaca (16 to 17.50°S, 68.5 to 70°W; 3810 m above sea level) occupies a portion of the northern Altiplano, a large internally drained plateau in the highlands of Bolivia and Peru (Figure 1). High mountains of the eastern and western cordillera of the Andes surround the lake. At present many of the highest peaks in the watershed are glaciated, although all of the glaciers that have been studied are currently in retreat (Francou *et al.*, 2003). The lake consists of a large (7,131 km²) main basin (Lago Grande, maximum depth 284 m, mean depth 125 m), which is connected by the Straits of Tiquina (25 m sill depth) to a smaller (1,428 km²) shallower basin

(Lago Huinamarca, maximum depth 42 m, mean depth 9 m). Contemporary hydrologic inputs are balanced between direct rainfall (47%) and inflow largely from six major rivers (53%). In the modern lake, water export is mainly via evaporation (~91%), with the remainder accounted for by variable discharge from the sole surface outlet of the lake, the Río Desaguadero (Roche *et al.*, 1992). This near balance between water input and water loss by evaporation indicates the potential sensitivity of Lake Titicaca and its sediment record to climate variability. The modern lake is oligosaline (0.1 g L⁻¹ salinity) and moderately productive (mesotrophic).

Modern mean annual surface-air temperature decreases from about 10 °C near Lake Titicaca to less than 4 °C in the southern Altiplano. Mean annual precipitation in the Lake Titicaca watershed varies from greater than 1100 mm in the center of Lago Grande; to about 700 mm at Puno, Peru, on the western shore of Lago Grande; 800 mm at Copacabana, Bolivia, on the southern shore of Lago Grande; and about 600 mm at El Alto (La Paz), Bolivia, to the southeast of the lake (data from SENAMHI, Peru and Bolivia). Precipitation exceeds poten-

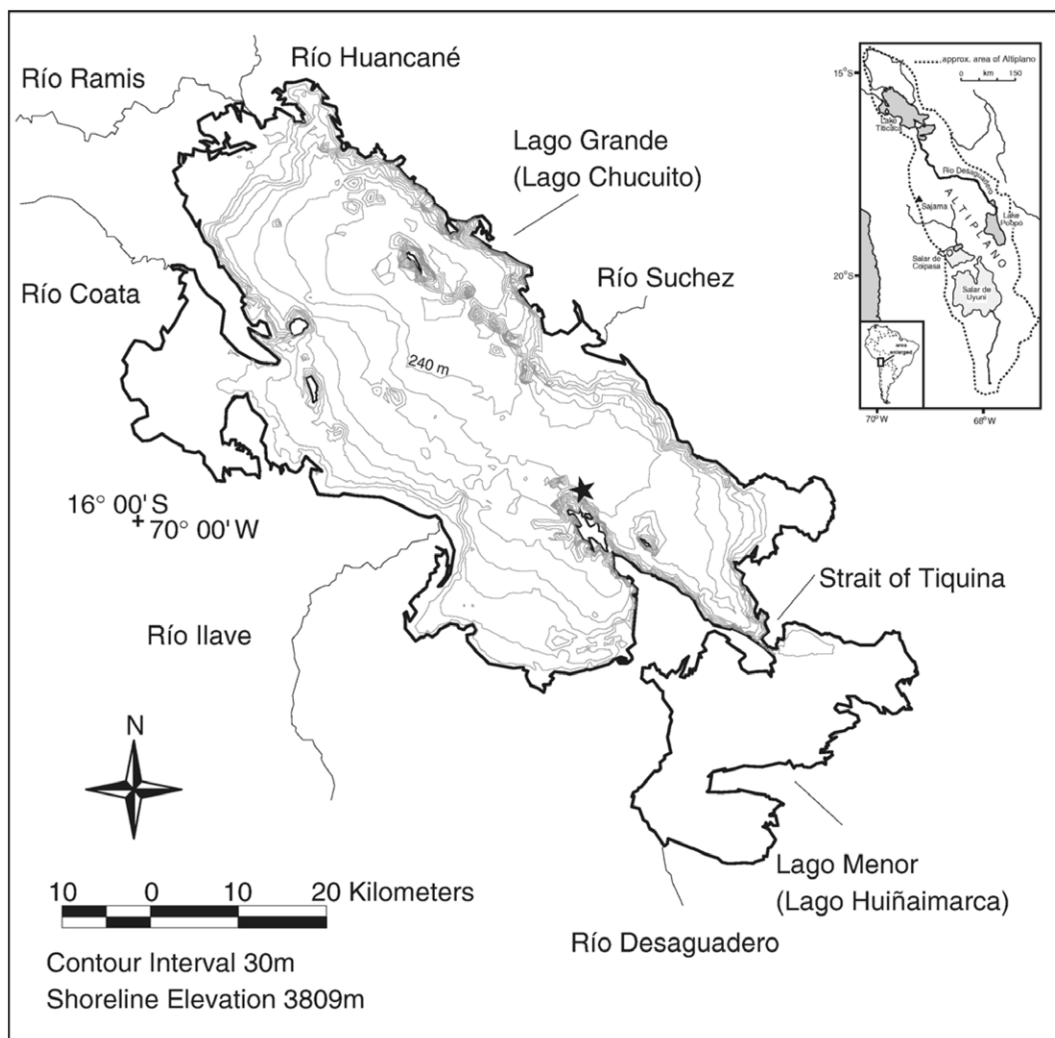


Figure 1. Map of Lake Titicaca, Bolivia/Peru, and the location site of core LT01-2B (star). Inset map shows the lake's location within the Altiplano (dotted line).

tial evapotranspiration, leading to a positive water balance for most of the Lake Titicaca watershed (Mariaca, 1985). On the northern Altiplano, about 80% of the precipitation occurs in the months of December through March and is largely a product of the South American summer monsoon (SASM) (Zhou and Lau, 1998). The nearly continent-wide scale of the circulation associated with the SASM implies that a record of climate limited to only one watershed nevertheless should be able to capture the major variations within this climate system. Most of the atmospheric moisture advected into the watershed of Lake Titicaca originates in the tropical Atlantic Ocean; a significant portion of this moisture is recycled by condensation and evapotranspiration along its trajectory across the Amazonian lowlands and the cloud forests of the eastern cordillera of the Andes (Salati *et al.*, 1979 and Vuille *et al.*, 2003a).

In the instrumental period, inter-annual to multi-decadal variability of northern Altiplano precipitation, which brings about large changes in lake level, is known to be forced by variations of tropical Pacific (Vuille *et al.*, 2003b and Vuille *et al.*, 2000) and tropical Atlantic (Hastenrath *et al.*, 2004) sea-surface temperature (SST). Both also have been shown to be important forcings of precipitation in climate (e.g. Cox *et al.*, 2004) and paleoclimate models (Cook and Vizy, 2006). On paleoclimatic time scales, a variety of factors are thought to have a role in forcing climate variability in tropical South America—these include orbitally-driven changes of seasonal insolation affecting precipitation in the SASM (Baker *et al.*, 2001a, Martin *et al.*, 1997, and Seltzer *et al.*, 2000); changes in climate boundary conditions, such as the presence or absence of large northern hemisphere ice sheets, CO₂ concentrations and global temperature (Garreaud *et al.*, 2003); anomalous SSTs in the tropical Pacific (Bradley *et al.*, 2003) or Atlantic (Vizy and Cook, 2005); and anomalous SST latitudinal gradients in the tropical Atlantic, related to migration of the inter-tropical convergence zone (Baker *et al.*, 2001b, Baker *et al.*, 2005, and Haug *et al.*, 2001). The resultant variability of climate changed the water balance and glacial mass balance on the northern Altiplano and was imprinted in the sedimentary record of Lake Titicaca.

Methods

Field

In May and June 2001, we raised overlapping drill cores from three different locations within Lake Titicaca, using the GLAD 800 drilling platform and coring system. Sediments were recovered using hydraulic piston coring above approximately 50 m below lake floor (mblf). Deeper firmer sediments were recovered using rotary drilling with an extended core barrel. Both coring technologies are similar to those employed by the Ocean Drilling Program. Here we report on analyses of a continuous core sequence from site LT01-2B (Figure 1) located to the east of Isla del Sol (235 m water depth), drilled to a total depth of 136 mblf. Magnetic susceptibility (MS) was measured in the field, and smear slides were made from core catcher sediments. The cores were shipped back to the U.S.

and are stored at the University of Minnesota Lacustrine Core Repository (LACCORE).

Laboratory

Continuous logging of MS, p-wave velocity, density and porosity were completed at LACCORE. Photographs and detailed sedimentological descriptions were made of the pair of overlapping cores from the site. Core 2B was subsampled at a resolution of 10 cm in units of apparently uniform lithology. In units of more variable lithology, including laminated sequences, subsampling of channel samples was done at continuous 2-cm intervals.

Organic carbon (TOC), %CaCO₃ and $\delta^{13}\text{C}$ of TOC were analyzed at high temporal resolution (1700–2500 samples of each). Samples were dried, powdered, weighed and leached in buffered (pH 5.5) ammonium acetate-acetic acid. Weight percent calcium carbonate was calculated from atomic absorption spectrophotometric (Perkin Elmer 5000) determination of dissolved calcium, assuming that all calcium was originally present as calcium carbonate. The acid-insoluble residue was rinsed several times in reagent-grade water. Portions of this residue were dried and weighed prior to determination of TOC and its stable isotopic composition. Stable carbon isotopic compositions were measured on a Finnigan MAT Delta Plus XL isotope mass spectrometer in the Duke University Environmental Stable Isotope Laboratory and are reported relative to the PDB standard. Reproducibility for replicate TOC analyses was better than 0.5%. The precision for the $\delta^{13}\text{C}$ measurements was $\pm 0.2\text{‰}$.

Diatom species composition was determined at 20-cm intervals (~700 samples) throughout the drill-core sequence. Samples for diatom analysis were treated with 10% hydrochloric acid and cold hydrogen peroxide to respectively remove carbonates and organic matter and then were rinsed to remove oxidation by-products. Prepared samples were dried onto coverslips, and the coverslips were mounted onto slides with Naphrax. Species were identified on a Zeiss Axioskop 2 microscope with a 1000 \times (N.A. = 1.40) oil immersion objective. At least 300 diatom valves were counted on each slide. Diatom abundance in each sample is expressed as a percent of the total diatom count. Diatom taxa are grouped into one of four main ecological groups (freshwater plankton, saline plankton, salinity indifferent plankton and benthic) based on known ecological affinities (Servant-Vildary, 1992 and Tapia *et al.*, 2003). The detailed stratigraphy of individual species will be published elsewhere.

Radiocarbon measurements were made on 18 acid-leached bulk organic carbon samples in the uppermost 26 m at site 2B. ^{14}C dates (Table 1) were calibrated using CALIB 4.4.2 for ages less than 20,000 ^{14}C yr before present (BP) (Stuiver *et al.*, 1998). For older sediments, we used the calibration curve of Hughen and coworkers (Hughen *et al.*, 2004). No reservoir correction was applied because surface sediments from box cores from the main basin of Lake Titicaca do not show a reservoir effect (Baker *et al.*, 2001b). Details associated with ra-

Table 1.

Radiocarbon ages as determined by accelerator mass spectrometry (AMS) dating of the total organic carbon fraction of the bulk sediments

Lab #	Drive-section-depth	Core depth (mblf)	$\delta^{13}\text{C}$ (per mil PDB)	^{14}C age (yr BP)	Age error (yr)	Calendar age (yr BP)	Upper age	Lower age
CURL-6087	1H-CC	0.108	-21.5	3050	30	3268	3164	3354
AA46943	2H-1 5-6 cm	0.28	-19.3	3625	43	3933	3782	4084
AA46944	2H-1 77-78 cm	1.00	-20.4	5988	52	6818	6672	6947
AA46945	2H-2 9-10 cm	1.83	-22.2	8601	62	9584	9474	9815
AA46946	2H-2 116-117 cm	2.90	-27.4	12,123	81	14,217	13,701	15,344
CURL-6088	2H-CC	3.12	-26.2	14,360	50	17,211	16,721	17,722
AA46947	3H-2 98-99 cm	5.715	-28	12,238	72	14,356	13,843	15,394
CURL-6089	3H-CC	6.135	-25.2	18,790	65	22,311	21,611	23,043
AA46948	4H-1 98-99 cm	7.21	-25.9	20,000	150	23,210		
CURL-6090	4H-CC	9.04	-24.3	20,760	80	24,046		
CURL-5941	5H-CC	11.84	-24.3	22,310	140	25,752		
CURL-5942	6H-CC	15.04	-24.64	24,190	160	27,821		
AA46949	7H-2 4.5-5.5 cm	16.78	-26.46	25,110	222	28,834		
AA46949	7H-2 4.5-5.5 cm	16.78	-1.52	25,310	220	29,054		
CURL-5943	7H-CC	18.04	-25.1	28,390	180	32,444		
CURL-5944	8H-CC	21.15	-25.25	33,370	200	37,925		
AA46950	9H-1 49-50 cm	21.72	-26.8	31,230	660	35,569		
CURL-5945	9H-CC	24.16	-25.98	36,680	270	41,142		
AA46951	10H-2 25-26 cm	25.99	-27.1	37,900	1900	41,416		
CURL-6091	10H-CC	27.09	-27	> 43,000				
CURL-6092	16H-2-50 cm	44.25	-19.8	> 43,000				
CURL-5946	16H-CC	45.23	-20.18	> 52,000				

Calendar age calibration of the ^{14}C ages is based on CALIB 4.4.2 (Stuiver *et al.*, 1998) for ages less than 20,000 ^{14}C yr before present (BP); calendar ages indicate the median probability and the 2-sigma upper and lower age limits. For older sediments, we used the calibration curve of Hughen (Hughen *et al.*, 2004). Core depths are in meters below the lake floor (mblf); CC is the core catcher at the base of a given drive.

diocarbon dating of Lake Titicaca sediments have been discussed previously (Rowe *et al.*, 2003).

Nine aragonite-rich layers from 3 core sections were subsampled for U/Th dating (Table 2) at the University of Minnesota. The mass of each subsample was kept to a minimum (9 to 60 mg) in order to subsample the most aragonite-rich portions of each layer, thereby minimizing the amount of thorium-rich detrital material. Each subsample was dissolved and spiked with a mixed ^{229}Th - ^{233}U - ^{236}U tracer (Cheng *et al.*, 2000). Uranium and thorium were separated using anion exchange techniques similar to ones described by Edwards *et al.* (Edwards *et al.*, 1987). The uranium and thorium isotopic compositions were determined on a magnetic sector inductively coupled plasma mass spectrometer (Finnigan Element) using the multiplier in ion-counting mode (Shen *et al.*, 2002).

Ages were calculated (Broecker, 1963) using half-life values for ^{230}Th and ^{234}U determined by Cheng *et al.* (2000) and the value for ^{238}U determined by Jaffey *et al.* (1971).

^{40}Ar - ^{39}Ar dating of interbedded ashes in the drill cores produced apparently unreliable ages, possibly because the samples contain detrital biotite that was not associated with the direct ash fall. We previously observed an Ar-Ar date $> 10^6$ yr on a 4-cm-thick ash from Lake Titicaca sediments that was bracketed by AMS ^{14}C dates of $22,200 \pm 90$ and $23,000 \pm 100$ ^{14}C yr BP. Thus, we do not report the Ar-Ar dates in this paper, and we do not utilize them in our age model.

Preliminary studies of the geomagnetic properties of the sediments (S. Brachfield, personal communication, 2006) indicate that the magnetic record was strongly affected by diagenesis, hence was not suitable for paleomagnetic dating.

Table 2. ^{230}Th dating results for samples from LT01-2B

Drive-section depth (cm)	$\delta^{234}\text{U}$ (mblf)	Depth (ppb)	^{238}U (ppb)	^{232}Th (ppm atomic)	$^{230}\text{Th}/^{232}\text{Th}$ (measured)	$\delta^{234}\text{U}$ (activity)	$^{230}\text{Th}/^{238}\text{U}$ (ka uncorrected)	^{230}Th Age (ka corrected)	^{230}Th Age (initial)
2H-1-26	0.49	1145 \pm 2	241.2 \pm 0.8	10.5 \pm 0.2	623.7 \pm 2.6	0.1337 \pm 0.0027	9.3 \pm 0.2	5.5 \pm 3.9	633.6 \pm 7.4
17H-1-17.5	45.404	6533 \pm 11	848 \pm 5	126.0 \pm 1.2	372.8 \pm 1.6	0.9901 \pm 0.0075	128.1 \pm 1.8	125.6 \pm 3.1	531.6 \pm 5.5
17H-1-20.5	45.434	4800 \pm 6	324 \pm 1	238.3 \pm 1.3	374.1 \pm 1.4	0.9727 \pm 0.0042	123.9 \pm 1.0	122.6 \pm 1.6	529.0 \pm 3.1
17H-1-100.5	46.232	4139 \pm 7	1168 \pm 7	57.5 \pm 0.6	374.6 \pm 1.8	0.9826 \pm 0.0089	126.1 \pm 2.1	120.6 \pm 6.0	526.6 \pm 9.2
17H-1-102.8	46.255	4416 \pm 7	623 \pm 4	113.3 \pm 1.2	361.9 \pm 1.7	0.9682 \pm 0.0087	125.1 \pm 2.0	122.3 \pm 3.4	511.3 \pm 5.5
17H-2-13	46.865	4229 \pm 10	1955 \pm 128	38.4 \pm 0.6	359.8 \pm 2.6	1.075 \pm 0.015	152.7 \pm 4.4	143 \pm 11	540 \pm 16
17H-2-40.5	47.139	5040 \pm 4	2306 \pm 22	38.2 \pm 0.6	404.1 \pm 1.2	1.058 \pm 0.014	138.0 \pm 3.3	129 \pm 10	582 \pm 16
17H-2-63.3	47.366	2631 \pm 4	1958 \pm 16	23.4 \pm 0.4	346.6 \pm 1.9	1.053 \pm 0.014	149.8 \pm 4.0	135 \pm 17	507 \pm 24
18H-1-31	48.536	3429 \pm 3	3001 \pm 27	23.4 \pm 0.4	355.3 \pm 1.2	1.242 \pm 0.018	215 \pm 8	198 \pm 20	622 \pm 35

The error is 2σ . $\lambda_{230} = 9.1577 \times 10^{-6} \text{ yr}^{-1}$, $\lambda_{234} = 2.8263 \times 10^{-6} \text{ yr}^{-1}$, $\lambda_{238} = 1.55125 \times 10^{-10} \text{ yr}^{-1}$. Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$.

Discussion

Core lithology

Sediments at site LT01-2B (Figure 2) primarily consist of two alternating lithologic units: unit 1, gray mud that has high MS, no carbonate and commonly has low TOC; and unit 2, tan to green-gray mud that is laminated or thinly bedded and contains carbonate, high TOC and low MS. The transitions between units 1 and 2 are variable in terms of the sequences of change in MS, TOC and carbonate content. Based on previous work in the basin (Baker *et al.*, 2001b and Seltzer *et al.*, 2002), we interpret unit 1 to be predominantly of glacial–fluvial origin, with a high detrital content derived from erosion of the surrounding cordillera. In contrast, unit 2 is primarily of autochthonous lacustrine origin. Unit 2 sediments were deposited during times when detrital input from the surrounding watershed was low. At the same time, lake level fell below the outlet, increasing the salinity of the lake and bringing about calcium carbonate supersaturation and its crystallization from the water column. Thus, this second lithology represents periods of reduced precipitation and reduced glacial extent. The 136-m sedimentary sequence at LT01-2B consists of four major cycles expressed by the interbedding of the two lithologic units (Figure 2). This stratigraphy indicates that the cored interval spans four major cycles of regional glacial expansion and retreat.

Core chronology

The chronology of the core (Figure 3) is constrained by radiocarbon measurements in the uppermost 25 m (Table 1) and by uranium series ages on discrete aragonite laminae that were relatively free of detrital sediments in the upper 48 m (Table 2). The ^{238}U concentrations of the aragonite subsamples were high (1.1. to 6.5 ppm), roughly comparable to concentrations in coralline aragonite and ideal for U/Th dating. ^{232}Th concentrations were low but significant, so that small corrections for initial ^{230}Th were necessary. Corrections were small enough so that the assumption of an initial bulk earth $^{230}\text{Th}/^{232}\text{Th}$ value $((4.4 \pm 2.2)(10^{-6}))$ was sufficient. We assumed an error of $\pm 50\%$ in this value. A corrected U-series age on an aragonite layer in the Holocene sediments (0.49 mblf) overlaps with the corresponding radiocarbon ages. The ages of the four samples in section 17H-1 are particularly robust, as the corrections for initial ^{230}Th in these samples are very small. Therefore, we have great confidence in averaging their ages and assigning a mean age of 122.8 ka to the mid-point of the depth interval (45.842 mblf). There is little doubt that the dated aragonite layers in 17H-1 correlate with the last interglacial period (MIS5e). In the construction of our age model, we reject the ages of the four deeper aragonite layers on the basis of their higher initial ^{230}Th values.

Comparison of the uppermost U-series ages and the ^{14}C chronology extrapolated to the base of the uppermost high MS

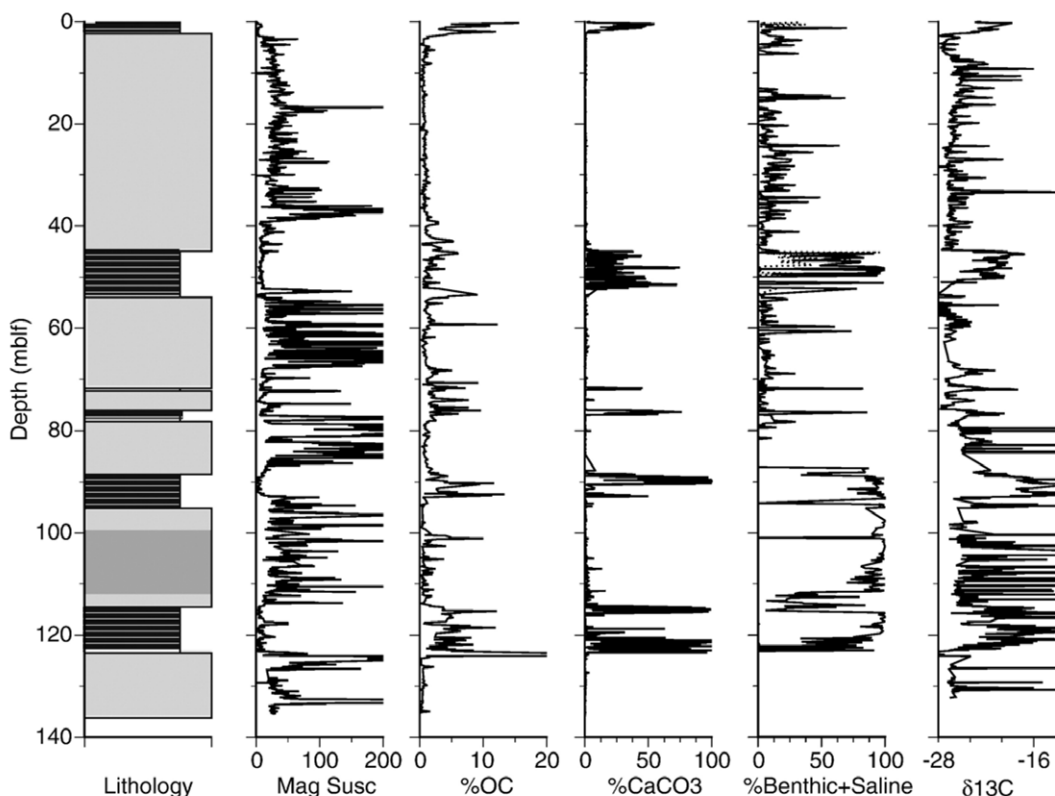


Figure 2. Stratigraphy of the LT01-2B drill core on a depth scale (meters below lake floor, mblf) including (a) lithology: black areas are carbonate-rich laminated sediments; gray areas are carbonate-poor silts and clay—the darker gray unit is rich in iron sulfides; (b) magnetic susceptibility (SI units), (c) % total organic carbon, (d) % calcium carbonate, (e) % benthic (solid line) and saline (dotted line) diatoms, (f) carbon isotopic value of bulk organic matter (per mil PDB).

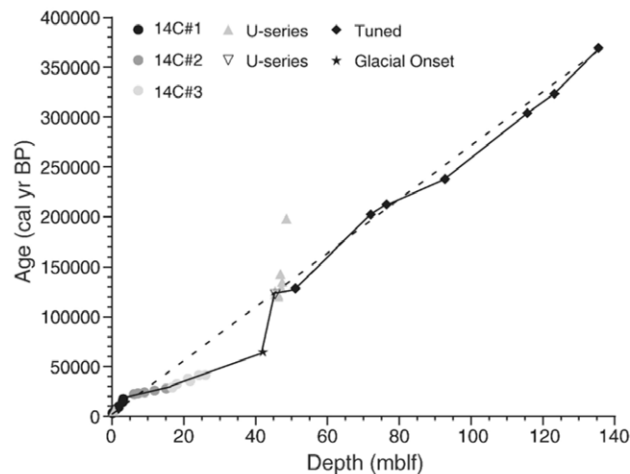


Figure 3. Data used in construction of a chronological model for the LT01-2B drill core, including AMS ^{14}C analysis of bulk organic matter, U-series dating of discrete aragonite laminae and tie points used in tuning the core calcium carbonate record to the Vostok CO_2 record. The calendar age–depth relationship of the ^{14}C dated portion of the core shows grouping of data points with three distinct slopes (Holocene plus late glacial, 20–30,000 yr BP, > 30,000 yr BP); separate equations were used to calculate calendar age from depth in each of these segments. The equation for sediments > 30,000 cal yr BP was extrapolated to the onset of the glacial age sediments at 42 mblf (marked by a star). U-series age measurements are shown with the filled triangles; the open triangle is the mean of the 4 ages used for the age model (see text). For other sections of the core, ages were interpolated between individual tie points derived from tuning the LT01-2B carbonate record to the Vostok CO_2 record (see Table 3).

lithologic unit suggests either slow or no sediment accumulation during MIS 5a through 5d. We assumed the former in modeling sediment accumulation rates because there is no clear vi-

sual or stratigraphic evidence for a hiatus at this depth in the core. In any case, in this portion of the record, we avoid making arguments that are dependent on the details of our age model.

In the upper 48 m of the core, high calcium carbonate concentrations, which represent times of lowered lake level, occurred during the mid-Holocene and MIS5e, thus during global warm periods. It is on the basis of this observation that we chose to tune the down-core peaks of high calcium carbonate (Figure 4) to global temperature maxima to create a tuned age model for sediments deposited prior to 122.8 ka (the mean of the four uppermost U-series ages in core section 17H-1). Two logical records of global temperature that can be used to create a tuned age model extend far enough back in the past. We chose the Vostok CO_2 record (Petit *et al.*, 1999) because it is believed to be a good proxy for global temperature (e.g. Shackleton, 2000) and it is relatively highly resolved. The alternative choice for a global temperature target is SPECMAP, but it has lower resolution, and the $\delta^{18}\text{O}$ record of marine foraminifera that comprise SPECMAP is a less reliable proxy for global temperature (Shackleton, 2000). It should be pointed out that the dating of the Vostok ice cores and SPECMAP (and similar derivatives) relies on orbital tuning to supply ages of various control points. Ages interpolated between control points rely on assumptions about constant sedimentation rates (for sediment cores) or glaciological model assumptions (for ice cores). Our age model, with the exception of ages established by radiocarbon or U/Th analyses, is based upon a similar methodology: assumption of the ages of the tie points that are ultimately derived from orbital tuning and interpolated assuming constant sedimentation rate between each tie point.

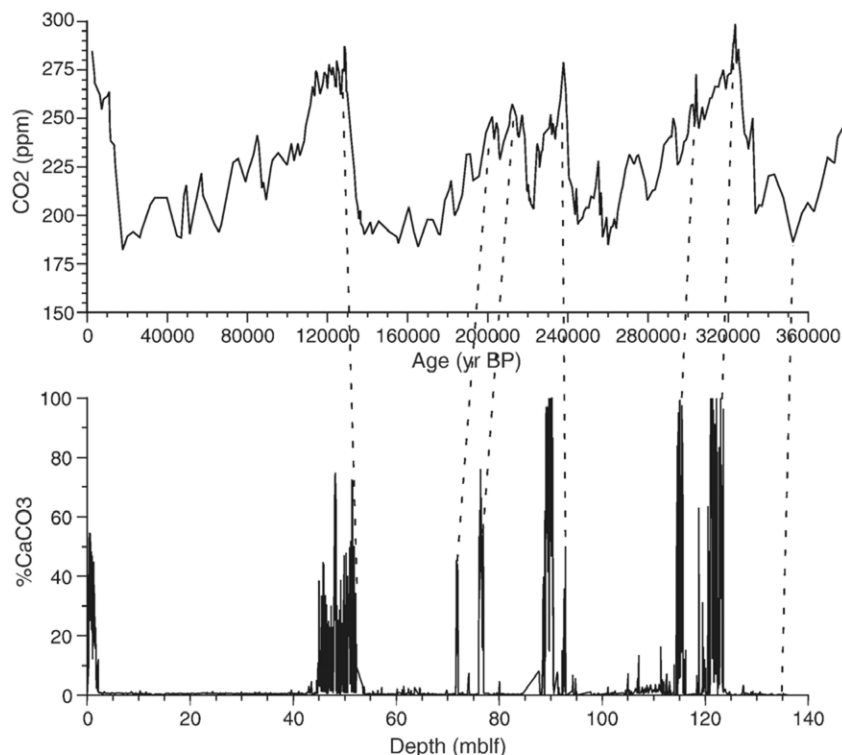


Figure 4. LT01-2B calcium carbonate concentration plotted on a depth scale showing tie points (dashed lines) for tuning the lake stratigraphy to the Vostok CO_2 record in order to derive an age model for basal core sediments (see text for further details).

The parsimony of our tuning approach is evident because the resultant chronology yields a nearly constant sediment accumulation rate over the entire core sequence (Figure 3) and because the age-depth model intersects the zero depth intercept at nearly zero age (350 yr BP). Application of this age model to the drill core sequence indicates that the 136-m core spans approximately the last 370,000 yr (Figure 5). A similar basal age for the core is obtained by summing the quotients of the two different mean sediment accumulation rates characteristic of unit 1 (1.0 mm yr^{-1}) and unit 2 (0.25 mm yr^{-1}) and the respective total thicknesses of both units. These characteristic sedimentation rates are those obtained respectively from the LGM and Holocene units in our ^{14}C -dated piston cores from Lake Titicaca (Baker *et al.*, 2001b). This concordance lends additional support to our proposed age model.

In summary, our age model consists of 11 sequential linear segments variously derived from ^{14}C , U-series and tuning methods (Table 3). Within each depth range, ages are interpolated assuming a constant sedimentation rate between tie points. Given the constraints due to imperfect chronology, our discussion below is qualified when referring to the earliest portion of the record.

Lake-level variation and its relationship to regional glacial cycles

Unit 1, characterized by high values of MS, marks the intervals of extensive glaciation in the cordillera surround-

ing Lake Titicaca (Seltzer *et al.*, 2002). The inference of lake level and salinity for the Lake Titicaca cores is based on calcium carbonate concentration, diatom stratigraphy and $\delta^{13}\text{C}$ isotopic measurements on sedimentary organic carbon, which have been established previously as reliable proxies in Lake Titicaca (Cross *et al.*, 2000, Rowe *et al.*, 2002 and Tapia *et al.*, 2003). During periods of high lake level, carbon derived from planktonic algal sources with lower $\delta^{13}\text{C}$ values dominates the organic carbon pool. Conversely, as lake level lowers, carbon derived from submersed littoral macrophytes having higher $\delta^{13}\text{C}$ values increases at the expense of carbon derived from

Table 3.
Equations used to calculate age (cal yr BP) for samples from LT01-2B

Upper depth (mblf)	Lower depth (mblf)	Age-depth equation, $x = \text{depth (mblf)}$
0	4.34	$4307x + 2540$
4.34	15.74	$615x + 18583$
15.74	41.98	$1410x + 6070$
41.98	45.429	$16602x - 631683$
45.429	51.102	$1023x + 76116$
51.102	72.02	$3529x - 51924$
72.02	76.446	$2275x + 38369$
76.446	92.71	$1571x + 92188$
92.71	115.64	$2884x - 29512$
115.64	123.213	$2579x + 5699$
123.213	135.45	$3766x - 140469$

See text for further details. [Corrected per QR 69 (2008), p. 342, corrigendum]

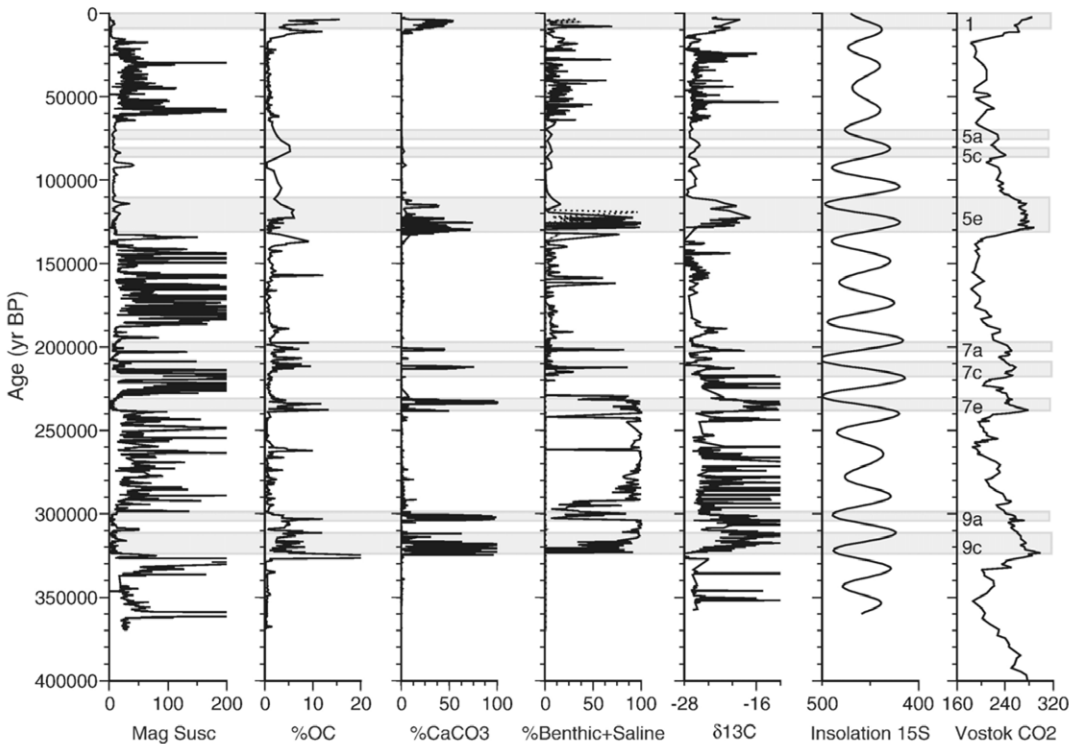


Figure 5. Stratigraphy of indicators in the LT01-2B drill core used to define intervals of glacial expansion in the cordillera and regional hydrology graphed on an age scale (cal yr BP): (a) magnetic susceptibility (SI units), (b) % total organic carbon, (c) % calcium carbonate, (d) % benthic (solid line) and saline (dotted line) diatoms and (e) carbon isotopic value of bulk organic matter (per mil PDB). Also shown are (f) January solar insolation (precession) at 15°S (Berger and Loutre, 1991) and (g) global warm and cool periods as indicated by the Vostok ice core CO_2 record (Petit *et al.*, 1999). Gray-shaded bands correspond to global warm intervals; the corresponding marine isotope stages (MIS) are labeled.

planktonic algae. Allochthonous carbon seems to be a relatively minor contributor to sedimentary organic carbon, probably because of the sparse vegetative cover of the watershed and the sediment-trapping effect of the abundant nearshore macrophytes. Diatom analysis differentiates between diatoms that grow in shallow regions of the lake (benthic) and species characteristic of deep open water (planktic), as well as species characteristic of freshwater or high salinity. These lacustrine proxies together demonstrate that increased glacial extent (positive glacial mass balance) in the cordillera surrounding Lake Titicaca almost always coincided with times when the lake was fresh (low calcium carbonate, high abundance of freshwater diatoms) and lake level was high (low abundance of benthic diatoms, more negative $\delta^{13}\text{C}$) indicating a positive water balance (Figure 2 and Figure 5). Conversely, reduced regional glaciation occurred during periods of negative water balance and shallow closed-basin conditions in the lake.

At issue is to what extent changes in hydrological and glacial mass balance were brought about by temperature (and other climatic) impacts on evaporation/sublimation rates or by precipitation variation. Blodgett and co-workers (1997) estimated that a 10 °C temperature decline unaccompanied by precipitation increase would be necessary to maintain the observed high lake stands on the Altiplano during the LGM. Given the much lower mean estimates of regional temperature change during the LGM (e.g. Bush *et al.*, 2004 and Cook and Vizy, 2006), it is likely that the observed history of regional glaciation and lake level in the southern tropical Andes was a product of both temperature and precipitation variation. Thus, the high lake levels and expanded cordilleran glaciers of the last glacial stage and prior glacial stages were brought about by a combination of lower temperatures and higher regional precipitation.

One possible exception to the correlation between high lake level and increased glaciation occurs during the basal glacial unit (Figure 2), when high percentages of benthic diatoms and high values of $\delta^{13}\text{C}$ coincide with a high-MS unit (101–111 mblf; estimated age of 262–292 ka). It may be that the lake was shallow during this time, although the low calcium carbonate concentrations and the diatom species composition suggest that it was fresh, not saline. Another possibility is that high river discharge into the lake associated with a cold wet climate transported littoral material into deep water during times of reduced pelagic production. Alternatively, it may be that the lake was deep and well stratified, with incomplete mixing over multiple years, which reduced nutrient availability to planktic algae but enhanced preservation of organic material, such as has occurred during some parts of the last century (Richerson *et al.*, 1986). Although the sediments in this interval are silts and clays, they are distinctive in that they are richer in iron sulfides and have slightly higher TOC values compared to the mean values of Unit 1 sediments and are thinly bedded to faintly laminated, characteristics consistent with the hypothesis of prolonged anoxia resulting from reduced water-column mixing.

Among the intervals of reduced glacial extent and low lake level, the upper portion of MIS5e (~44.8–52 mblf) was the most saline period in the LT01-2B record based on the high abundance of the saline diatom, *Chaetoceros muelleri*, cou-

pled with the high CaCO_3 concentrations and high $\delta^{13}\text{C}$ values (Figure 2 and Figure 5). This suggests a long period of closed-basin conditions, which allowed salinity to build up in the lake (Cross *et al.*, 2001). Pollen data for this interval show very high total pollen concentrations and high percentages of Chenopodiaceae/Amaranthaceae and are interpreted as indicative of extreme warmth and aridity (Hanselman *et al.*, 2005). In the northern tropical Andes, MIS5e was inferred to be warmer than other interglacials based on the high proportions of sub-Andean plant species (Van't Veer and Hooghiemstra, 2000).

The only other equally long paleohydrological record from the Southern Hemisphere tropics is from Salar de Uyuni (20°S 67°W) to the south of Lake Titicaca (Figure 1) (Baker *et al.*, 2001a and Placzek *et al.*, 2006). The three long-lived lacustrine intervals in the last 60 ka in the Salar de Uyuni drill-core record (Baker *et al.*, 2001a and Fritz *et al.*, 2004) are consistent with the inference of deep fresh conditions in Lake Titicaca during equivalent periods of time. The multiple short-lived lacustrine intervals of Salar de Uyuni between ~120 and 60 ka, however, occurred during an interval of either reduced or no sediment accumulation in the LT01-2B record. Because the Lake Titicaca record is poorly resolved in this interval, it is difficult to compare the two records. In the well-dated core from the Salar de Uyuni, MIS6 is characterized by salt deposition and dry conditions, in contrast to the inferred deep fresh conditions in Lake Titicaca (Figure 5). The difference suggests that the two basins are responding to different hydrological controls. It is possible that the 198-ka U-series age on aragonite at 48.55 m in the LT01-2B core (Table 2) is robust, and hence our age model is in error for the MIS6 sediments. We consider this unlikely, as we have argued above. A more probable explanation for the differing patterns between Salar de Uyuni and Lake Titicaca in the interval prior to 60 ka is that the two basins were not hydrologically connected at this time (Fritz *et al.*, 2004), possibly because this interval predates the down-cutting of the Rio Desaguadero, which presently connects Lake Titicaca with the southern Altiplano.

Timing of expanded glaciation in the Lake Titicaca watershed

The magnetic susceptibility increase at 42 mblf (Figure 2) likely marks the onset of the most recent period of increased regional glaciation. Extrapolation of the radiocarbon chronology to this depth suggests that glacial ice expansion in the region began approximately 60,000 ^{14}C yr BP, thus during the latter part of MIS4. Both calcium carbonate concentration and benthic diatom abundance decrease prior to the MS increase, implying that lake-level rise, perhaps due to increased precipitation, preceded regional glacial advance.

Our age model for the lower part of the core suggests that prior to 60 ka, sustained intervals of increased magnetic susceptibility and thus expanded regional glaciation occurred in the periods 370–324, 300–238, 230–213 and 188–132 ka (Figure 5). Using ^{10}Be exposure ages and estimating rates of boulder erosion and surface uplift, Smith and co-workers (2005a) tentatively identified four periods of increased glaciation in the Lake Junin basin of Peru (10°S): > 425, 320–200, 170–125

and 85–21 ka. These intervals represent the ages of deposition or of exhumation of glacial moraines that were persistent and not overridden by subsequent glacier advance and, as such, are not equivalent to the continuous record of glacier erosion and transport deduced from the Lake Titicaca MS record. Nevertheless, there is reasonable accord between the data of Smith and co-workers and our own estimation of the ages of regional glaciation derived from the Lake Titicaca drill core.

Climate forcing

The penultimate low stand of Lake Titicaca is dated by U-series measurements to MIS5e. Our orbitally tuned time scale implies that earlier low stands of Lake Titicaca also correspond to global warm periods evident in the Vostok ice core (MIS 7a, 7c, 7e, 9a, 9c). Thus, the major wet-dry cycles seen in the Lake Titicaca drill core and the major cold-warm cycles of the Vostok sequence share an eccentricity period.

We believe that this warm-dry (or cold-wet) correspondence is due to the local increase of evaporation and decrease of water balance in Lake Titicaca due to global (and local) temperature increase (e.g. Blodgett *et al.*, 1997) as was argued previously for the record from the Sabana de Bogata in Colombia (Van't Veer and Hooghiemstra, 2000). A second cause for the relationship between higher temperatures and lower lake levels is climatic. Today, during ENSO warm events, atmospheric subsidence (and other dynamic and thermodynamic changes) suppresses precipitation over much of tropical South America (e.g. Zhou and Lau, 2001, their Figure 4), including the Altiplano—thus globally warm temperatures (during ENSO warm events) are often associated with dry conditions at Lake Titicaca on interannual time scales. Bradley and co-workers (2003) have suggested that such a relationship also may prevail on longer time scales. Furthermore, tropical Atlantic sea-surface temperatures (SST) may exert a similar climate control: above-average SSTs in the northern tropical Atlantic often are associated with dry conditions in the Amazon and on the Altiplano (e.g. Nobre and Shukla, 1996, Baker *et al.*, 2001b, and Baker *et al.*, 2005).

On somewhat shorter time scales, the importance of precession and its characteristic 20 kyr pacing in forcing the tropical monsoons has been demonstrated previously by climate modeling (Kutzbach, 1981) and in empirical studies in South America (Baker *et al.*, 2001a, Bush *et al.*, 2002, Cruz *et al.*, 2005, Hooghiemstra *et al.*, 1993, Martin *et al.*, 1997, and Wang *et al.*, 2004). But the penultimate low stand of Lake Titicaca, rather than dating to the last summer solar minimum (~ 32 ka), is coincident with MIS5e. Overall the relationship between precession and lake level is variable through the drill core sequence, both in core sections dated by ^{14}C and U-series dating, as well as in sections dated by tuning (Figure 5). Well-dated speleothem records from Brazil (Cruz *et al.*, 2005 and Wang *et al.*, 2004) indicate a strong relationship between precession and precipitation from the present to 140 ka, but this relationship apparently breaks down prior to 140 ka (Wang *et al.*, 2004) during global glacial period MIS6. A similar pattern was also observed at Salar de Uyuni (Fritz *et al.*, 2004). The lack

of a consistent relationship between precession and hydrologic variation on these long time scales speaks to the multiplicity of factors that are likely to influence precipitation variation, among them insolation, global ice volume and Pacific (Garreaud *et al.*, 2003) and Atlantic SSTs (Baker *et al.*, 2005). Thus, the impact of precession forcing relative to other climate drivers may have been dampened at various times, such as when the amplitude of insolation variation was relatively low.

Conclusions

The Lake Titicaca drill-core sequence represents the longest continuous record of both glaciation and hydrologic variation from the Southern Hemisphere tropics of South America and clearly documents four regional glacial–interglacial cycles in the tropical Andes. The general correspondence of regional glacial periods with intervals of high lake level and of periods of reduced glaciation with times of low lake level indicates that cold–wet and warm–dry conditions are the two (of the possible four) regional norms. Within the limits of our age model, the periods of regional glacial advance coincide with global glacial stages and periods of regional glacial retreat coincide with global interglacial stages. Thus, in the southern tropical Andes, climate conditions during global glacial stages are inferred to be cold and wet. This contrasts with the iconic drill core records from the northern tropics of South America (Hooghiemstra *et al.*, 1993) where global glacial stages are inferred to be cold and dry (a relationship to be expected for the northern subtropics given the opposite phasing of the precessional forcing relative to the southern subtropics). The nature of the relationship between the timing of regional glacial and hydrological cycles and the pacing of insolation variation and Northern Hemisphere glacial cycles suggests that regional water and glacial mass balances on the Altiplano were strongly influenced by both global-scale glacial boundary conditions (specifically, global temperature) and by precessionally paced, regional insolation forcing of the SASM.

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